Recent Trends in Land Surface Temperature on the Tibetan Plateau

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ABSTRACT

The diurnal, seasonal, and interannual variations in land surface temperature (LST) on the Tibetan Plateau from 1996 to 2002 are analyzed using the hourly LST dataset obtained by Japanese Geostationary Meteorological Satellite 5 (GMS-5) observations. Comparing LST retrieved from GMS-5 with independent precipitation amount data demonstrates the consistent and complementary relationship between them. The results indicate an increase in the LST over this period. The daily minimum has risen faster than the daily maximum, resulting in a narrowing of the diurnal range of LST. This is in agreement with the observed trends in both global and plateau near-surface air temperature. Since the near-surface air temperature is mainly controlled by LST, this result ensures a warming trend in near-surface air temperature.

1. Introduction

The Tibetan Plateau plays an important role in global climate and atmospheric circulation through orographic and thermal forcing mechanisms (e.g., Ye and Gao 1979; Ye 1981; Ye and Wu 1998). The ground surface of the Tibetan Plateau reaches elevations of greater than 4000 m above sea level, blocking the tropospheric circulation and impacting on the upper airflow across the Eurasian continent. The plateau surface absorbs the larger amount of incoming solar energy in daytime than its surrounding area, so that it directly heats the middle of troposphere above it. Therefore, it is essential to study energy balance between land surface and atmosphere over the plateau. The quantitative estimation of thermal effect is required to understand its influence on regional and global climate.

Land surface temperature (LST) is an important parameter for the monitoring of the energy exchange between the land surface and the atmosphere in terms of the sensible and latent heat fluxes. These fluxes are important when discussing the thermal effects of the Tibetan Plateau on the regional and global climate. Sensible heat flux is determined by temperature difference between the land surface and the air immediately above it. The fluctuation of air temperature is smaller than that of the LST on seasonal and interannual scales, and the variation in the temperature difference between the air and the surface varies depending mainly on the LST variation. Therefore, LST is a suitable quantity/parameter for analysis of the thermodynamic processes from the ground surface to the atmosphere. LST is also related to climatic variations caused by thermodynamic forcing. Therefore, research on the variation in the LST on the Tibetan Plateau is of climatological and meteorological significance.

The Intergovernmental Panel on Climate Change (IPCC) has pointed out the urgent need for the inclusion of long-term remote sensing–based LST data in global warming studies (Houghton et al. 2001). Satellite remote sensing offers the possibility of determining regional distributions of LST, but at present, long-term surface temperature datasets are only available for the
oceans. Surface temperature sensing for land areas has proved to be difficult due to the heterogeneities of the surface. An early effort was made by applying the National Oceanic and Atmospheric Administration (NOAA)/Advanced Very High Resolution Radiometer (AVHRR) sea surface temperature split-window technique to homogeneous land (e.g., Sobrino et al. 1996; Li and Becker 1993). Although the polar-orbiting satellite NOAA provides high spatial resolution, the measurement frequency is approximately 2 times per day, which is insufficient for diurnal cycle retrieval. According to surface observations during the Global Energy and Water Cycle Experiment (GEWEX) program for the Tibetan Plateau, the GEWEX Asian Monsoon Experiment/Tibetan Plateau (GAME/Tibet), it has shown that the LST has large diurnal range as well as the annual range (Fig. 1). This suggests the necessity to estimate the LST at shorter time intervals than twice daily. The diurnal cycle of LST is an important element of the climate system and can be obtained using a geostationary satellite. Whereas the polar-orbiting satellite provides high-spatial-resolution data, the geostationary satellite provides excellent temporal sampling over large parts of the globe from the equator to the middle latitudes. For the Tibetan Plateau, the Japanese Geostationary Meteorological Satellite 5 (GMS-5) is available for recording the LST distribution.

2. Algorithm and data

GMS-5 was launched on 18 March 1995 and provides coverage of the Asia–Pacific region from Hawaii to India. Continuous hourly information was recorded for about 8 yr to 22 May 2003 when operation was terminated. GMS-5 had an onboard visible and infrared spin scan radiometer (VISSR), which senses three infrared channels: two split-window channels (IR1, 11 μm; IR2, 12 μm) and a water vapor channel (6.7 μm). To obtain the LST distribution over the Tibetan Plateau, the split-window technique is applied to the radiances of these three infrared channels (Oku and Ishikawa 2004). This technique utilizes the difference in atmospheric absorption at two different wavelengths (11 and 12 μm) in a radiative transfer equation. LST, $T_{\text{sfc}}$, is determined as follows:

$$T_{\text{sfc}} = T_{\text{IR1}} + A(T_{\text{IR1}} - T_{\text{IR2}}) - B - C(1 - \epsilon) - D\Delta\epsilon,$$

where $T_{\text{IR1}}$ and $T_{\text{IR2}}$ are GMS-5 brightness temperatures at 11 and 12 μm. Here, $\epsilon = (e_1 + e_2)/2$ is the average emissivity over both channels, and $\Delta\epsilon = (e_1 - e_2)$ is the spectral variation in emissivity. Surface emissivity mainly depends on the vegetation, so that diurnal variations are expected to be relatively smaller than seasonal variations. According to Sobrino and Raisouni (2000), $\epsilon$ and $\Delta\epsilon$ are estimated from the normalized difference vegetation index (NDVI) that is calculated from visible split-window channels of NOAA/AVHRR images. NDVI is computed as

$$\text{NDVI} = \frac{\rho_2 - \rho_1}{\rho_2 + \rho_1},$$

where $\rho_1$ and $\rho_2$ are the reflectance values yielded by AVHRR channel 1 and channel 2, respectively. NDVI is estimated as 1-month mean value. The coefficients $A$ through $D$ in Eq. (1) are given by

$$A = \frac{1 - \tau_1}{\tau_1 + \tau_2},$$

$$B = A(1 - \tau_2)(T_{\text{1air}} - T_{\text{2air}}),$$

$$C = \frac{1 - \tau_1 \tau_3}{\tau_1 - \tau_2} \left( T_{\text{IR1}} - T_{\text{IR2}} \right) + \tau_3 \frac{T_{\text{IR1}}}{4.667},$$

$$D = \tau_2 AC,$$

where $\tau$ is total atmospheric path transmittance, $T_{\text{1air}}$ is the mean temperature of the atmosphere between the surface and top of the atmosphere. Subscripts 1 and 2 of $\tau$ and $T_{\text{air}}$ represent the 11- and 12-μm channels, respectively; $\tau_3$ is the transmittance at a zenith angle of 53°; and $\tau$ and $T_{\text{air}}$ vary dependently on satellite zenith
angle and precipitable water along the optical path. Precipitable water is estimated directly from the brightness temperature of the water vapor channel. As the weighting function of the 6.7-μm sensor has a maximum near 400 hPa and does not reflect humidity at lower levels, the 6.7-μm brightness temperature is usually a poor indicator of precipitable water. Over the Tibetan Plateau, however, the ground surface is in 500-600-hPa level, so that 6.7-μm brightness temperature is expected to be related to the precipitable water information there.

Cloud removal has an important part in the surface temperature retrieval process. To identify convective cloud activity, many researchers use satellite infrared measurements with a fixed threshold technique. But in this study, it is necessary to remove not only convective clouds but also all kinds of clouds. For this purpose, a variable threshold technique proposed in Oku and Ishikawa (2004) is used, that is,

\[ T_{IR1} < T_{IR1}^{*}(DOY,UTC) : \text{cloudy} \]
\[ T_{IR1} \geq T_{IR1}^{*}(DOY,UTC) : \text{cloud free}, \]

where the threshold \( T_{IR1}^{*} \) varies both seasonally [day of year (DOY)] and diurnally (UTC), and the values are determined on the basis of surface observations. As a result of adoption of this technique, it becomes possible to remove relatively warmer clouds in summer and detect colder ground surfaces in winter nighttime.

All available GMS/VISSR and NOAA/AVHRR images from 1996 to 2002 were prepared to estimate LST and NDVI distribution over the Tibetan Plateau. The study area is defined as the area higher than 4000 m. The data, in grid format over longitude and latitude with a resolution of 0.1°, were interpolated from the original image data and archived in line–pixel format.

LST varies dependently on land surface condition, such as vegetation coverage and soil moisture. NDVI is the most commonly used in vegetation change detection, vegetation classification, and so on. Lambin and Ehrlich (1995) have reported that the combination of LST and NDVI gives the better land-cover classification, and Lambin and Ehrlich (1996) have validated this method by comparing it with biophysically analyzed land-cover change.

The reliability of LST should be demonstrated by validation with independent datasets. Monthly, the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) data (Xie and Arkin 1996, 1997) and the Special Sensor Microwave Imager (SSM/I) monthly snow-cover product (Basist et al. 1996, 1998) are used to compare their variations with LST.

The intensive observation period and long-term observation of GAME/Tibet and following coordinated enhanced observing period (CEOP) Asia–Australia monsoon project (CAMP) on the Tibetan Plateau (CAMP/Tibet) have been done successfully in the past 7 yr. LST data measured by infrared thermometer in automatic weather stations (AWSs) at D66, MS3608, and Gaize from 1997 to 2003 are used for verification and comparison with those retrieved from satellite. The infrared thermometers (OPTEX Thermo-hunter IK3 at Gaize, and CHINO Infrared Radiation Thermometer HR1-FL at D66 and MS3608) were installed to AWS at 1-m height above the ground. A datalogger sampled the LST data every second and the 10-min averages were recorded in it. The soil moisture observation data at the depth of 4 cm in MS3608 are used to compare with LST data. The observation interval was 1 h and the data were collected automatically. The soil moisture content measured by the time-domain reflectometer (TDR) probes refers to volumetric liquid water content when soil is thawed and to volumetric unfrozen water content when soil is frozen. The locations of observation points and land-use map (Dickinson et al. 1986) over the Tibetan Plateau are shown in Fig. 2.

The detailed model evaluation has been done and presented in Oku and Ishikawa (2004). The results of comparing estimated surface temperature from GMS-5 data using this algorithm with in situ surface measurements by AWS shows high correlation coefficient: it is nearly 0.9, with root-mean-square error (rmse) of about 10 K.

3. Results

Monthly averages of the daily mean LST were calculated from the available data for each grid, and the area average of the Tibetan Plateau was computed from grid values for each month from 1996 to 2002. About 60 000 hourly images were processed. It should be noted that the estimation of LST is only possible in cloud-free
areas. The ratio of cloud-free area to total grid was 51.5% over these 7 yr.

The interannual variation in the monthly mean LST recorded by GMS-5 is shown in Fig. 3 as anomalies from average years. To show the relative significance of the plots, the ratio of cloud-free pixels to the total is indicated by different marks. Daily mean LSTs in winter 1997/98 and 1999/2000 are relatively colder than other winters during these 7 yr. Monthly mean snow-cover fraction averaged across the Tibetan Plateau derived from SSM/I in these winters was larger than other winters (Fig. 4), and Sato (2001) reported that the days of snow cover in the winter 1997/98 was the longest from 1993 to 1998 at the meteorological stations in Naqu [World Meteorological Organization (WMO) ID 55299, 31.5°N, 92.0°E, 4508 m], Sogxian (ID 56106, 31.9°N, 93.8°E, 4024 m), and Qamdo (ID 56137, 31.2°N, 97.2°E, 3307 m). Additionally, monthly mean air temperatures measured routinely at the Tuotuohe meteorological station (ID 56004, 34.2°N, 92.4°E, 4535 m) from 1990 to 2000 indicate colder winters in 1997/98 and 1999/2000 than others (Zhang et al. 2003). It is readily conceivable that monthly mean daily land surface temperature drops to a lower value in the winter that has heavier snowfalls or colder air temperature than other winters.

The ground surface over the Tibetan Plateau is very dry before onset of the Asian monsoon, April to May. Incoming solar radiation in this period becomes stronger than that in winter season, and the cloud amount is less than that in summer. Consequently, the ground surface is easily heated in daytime and cooled in nighttime, so that diurnal range of LST becomes the largest of the year. After the onset of monsoon, surface soil moisture increases gradually because of precipitation. Reflecting this, the diurnal range of LST drops. And diurnal range of LST becomes its annual minimum in midmonsoon season.

Dai et al. (1997) suggest that precipitation amount is a good candidate to explain the decrease in diurnal range of near-surface air temperature. Similarly, the decrease in diurnal range of LST can be explained by the analysis of variation in precipitation amounts. Figure 5 shows interannual variation of monthly precipitation amount averaged across the plateau. In the summers of 1998, 2000, 2001, and 2002, precipitation amount is larger than that in other summers. Anomalies in diurnal variation of LST indicate negative high value in these summers. The increased precipitation may have contributed to the LST diurnal range decrease through intensification of surface evaporative cooling in daytime and suppression of radiative cooling in nighttime. The correlation between diurnal range of LST and precipitation is expected to be substantially negative value. Normalized monthly precipitation amount anomaly is compared with the corresponding
normalized monthly mean LST diurnal range anomaly and is shown in the left panel of Fig. 6. The correlation coefficient between both anomalies is $\rho = -0.435$.

The increased precipitation can bring about the increase in soil moisture there, and soil moisture can damp diurnal range of LST through evaporative cooling. This is especially effective during the day when the planetary boundary layer is unstable and the potential evapotranspiration is high. Interannual variation of soil moisture is compared with that of LST diurnal range by using surface observation data measured at MS3608, which has the longest and steadiest record. The right panel of Fig. 6 shows the relationship between soil moisture and diurnal range of LST. The correlation coefficient is $\rho = -0.470$. Although the negative correlation is not strong, these results support the consistent and complementary relationship between LST and precipitation on a monthly basis.

From 1996 to 2002, daily mean LST (upper middle panel in Fig. 3) is rising slightly. The trend obtained by linear regression is $0.1216$ K decade$^{-1}$. Daily minimum LST (lower middle panel) is rising $0.2865$ K decade$^{-1}$, whereas the maximum LST (top panel) is $0.0049$ K decade$^{-1}$. Since daily maximum LST has nearly no trend, diurnal range is decreasing gradually at the rate of $0.0745$ K decade$^{-1}$ (bottom panel). Similar trends are also detected in surface observation data measured by AWS at MS3608 from 1997 to 2003, as shown in Fig. 7 and Table 1.

To investigate the horizontal distributions, LST trends are calculated for each grid and then plotted in panel in Fig. 3) is rising slightly. The trend obtained by linear regression is $0.1216$ K decade$^{-1}$. Daily minimum LST (lower middle panel) is rising $0.2865$ K decade$^{-1}$, whereas the maximum LST (top panel) is $0.0049$ K decade$^{-1}$. Since daily maximum LST has nearly no trend, diurnal range is decreasing gradually at the rate of $0.0745$ K decade$^{-1}$ (bottom panel). Similar trends are also detected in surface observation data measured by AWS at MS3608 from 1997 to 2003, as shown in Fig. 7 and Table 1.

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<tr>
<td>Total plateau</td>
<td>Vegetated</td>
<td>Unvegetated</td>
<td>MS3608 (AWS)</td>
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<tr>
<td>Daily mean</td>
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<td>0.3529</td>
<td>0.0666</td>
</tr>
<tr>
<td>Daily maximum</td>
<td>0.0049</td>
<td>0.7698</td>
<td>$-0.2469$</td>
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<tr>
<td>Daily minimum</td>
<td>0.2865</td>
<td>0.3291</td>
<td>0.3344</td>
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<td>Diurnal range</td>
<td>$-0.0745$</td>
<td>0.7307</td>
<td>$-0.3666$</td>
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Fig. 8. Daily mean LST is increasing in almost all parts of the plateau, and daily minimum LST is increasing in the whole plateau. By contrast, the trend of daily maximum LST shows considerable regional variation. In the eastern and southwestern parts of the plateau, daily maximum LST is increasing, but decreasing in the northwestern and southern parts of the plateau. Reflecting this maximum LST trend pattern, the diurnal range of LST is decreasing in the northwestern and southern parts of the plateau. Decreasing trend in diurnal range of LST is found in more than one-half of the plateau. This diurnal range trend pattern corresponds with the pattern of seasonal variation in NDVI (see Fig. 9). NDVI values in the area classified as shortgrass are shown in Fig. 2 are increasing to summer. In semidesert/desert area, however, NDVI has almost no seasonal variation and keeps low value less than 0.1.

We now divide the plateau into two areas, vegetated (shortgrass) and unvegetated (semidesert and desert) area, based on Fig. 2, and then calculate monthly averages of NDVI and daily mean, maximum, minimum LST, and diurnal range of LST in each area.

Figure 10 shows monthly averages of daily maximum (left top), mean (left upper middle), minimum LST (left lower middle), and diurnal range of LST (left bottom) across the vegetated area. The corresponding values for the unvegetated area are shown in the right panels of Fig. 10. The slopes of the linear regressions are displayed in Table 1. Generally, high NDVI values correspond to more vegetation cover than lower values. The ground surface covered by vegetation canopy can retain water. When water is abundant, evaporation capacity of canopy becomes large, which controls amplitude in the diurnal range of LST across the vegetated area, especially during the monsoon period. Annual range of daily mean LST across the unvegetated area is larger than that in the vegetated area. During the monsoon season, the diurnal range of LST across the unvegetated area is about 5 K larger than that in the vegetated area (not shown).

Trends in daily mean LST across both areas are positive. The trend in daily maximum LST is 0.7698 K decade$^{-1}$ across the vegetated area, whereas it is $-0.2469$ K decade$^{-1}$ across the unvegetated area. The trend in daily minimum LST is 0.3291 K decade$^{-1}$ across the vegetated area, whereas it is 0.3344 K decade$^{-1}$ across the unvegetated area. Thus, the diurnal range of LST across the unvegetated area is decreasing (rate is $-0.3666$ K decade$^{-1}$), whereas that across the vegetated area is increasing (0.7307 K decade$^{-1}$).

The quantitative relationship between LST and NDVI for each area is depicted in Fig. 11. It is obvious that difference of vegetation-cover categories gives different slope in the LST/NDVI diagram. Annual range of LST in the vegetated area is smaller than that in the

![Fig. 8. Horizontal distributions of the trend obtained by linear regression at each grid. (upper left) Daily mean, (lower left) minimum, and (lower right) maximum LST, and (upper right) diurnal range of LST.](image1)

![Fig. 9. The 7-yr-mean value NDVI across the Tibetan Plateau in (left) January and (right) July.](image2)
unvegetated area, which is due to relatively high NDVI values during the monsoon period. The interannual variation trajectories are plotted as white circles for 1996 as the beginning year of these 7 yr, and as large black dots for 2002 as the end year. NDVI across the unvegetated area in 2002 is the smallest for these 7 yr. The diurnal range of LST in the unvegetated area seems to be decreasing. No clear LST and NDVI intra-annual variation is seen in the vegetated area.

4. Conclusions and remarks

A long-term LST diurnal cycle dataset was developed from GMS-5 observations. This dataset covers cloud-free areas on the Tibetan Plateau and spans from 1996 to 2002. As presented, daily mean, maximum, minimum LSTs, and diurnal range of LST were analyzed and the following features were revealed: 1) The coldest monthly mean LST in the winter 1997/98 is in excellent agreement with the largest snow-cover fraction averaged across the Tibetan Plateau. 2) There is negative correlation between interannual variation of LST diurnal range and precipitation on monthly basis. 3) The daily minimum LST has risen faster than the daily maximum, resulting in a decreasing of the LST diurnal range.

The increase in LST over the Tibetan Plateau is closely related to the increase of near-surface air temperature. The increase of near-surface air temperature has involved a faster rise in daily minimum than daily maximum temperature in many continental regions, and the diurnal temperature range decreases in many parts of the world (e.g., Houghton et al. 2001; Easterling et al. 1997). For the Tibetan Plateau, ground-based observation from 1951 to 1990 shows that the daily maximum near-surface air temperature is increasing slightly but the minimum is increasing significantly (Zhai and Ren 1999).

The increase of daily minimum near-surface air temperature reflects the evidence of enhancement of greenhouse effect, and the changes of both maximum and minimum near-surface air temperatures are mainly related to sunshine duration and atmospheric water vapor content (Zhai and Ren 1999). Additionally, these
asymmetric trends are partially related to increases in cloud cover (Karl et al. 1993). But there is no conclusive discussion to explain this asymmetry in the trends of both maximum and minimum near-surface air temperatures. Rising trend in daily mean LST could be attributed to intensification of surface heating due to increasing incoming solar radiation or reduction of radiative cooling from ground surface to the atmosphere. The fact that rising ratio of daily minimum LST is greater than that of maximum LST might support the latter possibility. It can be brought by frequency in cloud cover or precipitation, wet condition in ground soil, increasing water vapor in the near-surface air, and so on. Houghton et al. (2001) and many researchers found the increase in near-surface air temperature from ground-based observation data. The present result of LST is consistent with their work, and LST trends are also obtained from satellite data, which are superior to ground-based measurements in spatial representativeness. Since the near-surface air temperature is mainly controlled by LST, our result ensures a warming trend in near-surface air temperature.

The increasing trend in LST also suggests the increase in sensible heating from ground surface to the atmosphere. This increase might affect the formation of the Asian monsoon system and hence global climate. Although the precise cause of recent trends of LST over the Tibetan Plateau is yet to be known, they are in agreement with the observed trends in global near-surface air temperature.

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